ESTIMATING SNOWPACKS IN A DYNAMIC PRAIRIE ENVIRONMENT

H. Steppuhn[1]

ABSTRACT

The Northern Great Plains and the Canadian Prairies cover a vast area east of the North American Rocky Mountains. The climates of these plains and prairies exemplify the classic mountain "rain-shadows" at mid-latitudes with two dominant seasons: winter, and summer. Snow can blanket these lands at any time from September through June. The snowfall can accumulate on the ground forming snowpacks which vary widely in magnitude, areal coverage, and persistence from location to location and from year to year.

Widely varying snowpack depths (d), water equivalents (WE), and areal covers (A) typify cold, windy, prairie environments on the North American plains and prairies. The wide range in seasonal snowpack accumulations stems from the region's size and dynamic weather. Extremes are expected almost every year but typically occur in different locations. Although accurate estimates of snowpack d, WE, and A serve for effective resource management, they are often difficult to obtain. A technique, based on stratification according to terrain and vegetation, utilizes precipitation gauge accumulations to estimate mean snowcover WE-values as the season progresses. Gauge accumulations are adjusted for evaporation and meltwater releases by incorporating common meteorological station measurements. A model of the technique applied to data from Saskatchewan, improved r^2 values from 0.40 to 0.76 for regressions of precipitation gauge WEs with areal mean WEs measured in field snow surveys.

INTRODUCTION

The wise management of snowpacks depends on knowledge derived from accurate, quantitative measurements and surveys of the snow resource. Measurements of snowpack water equivalent, depth, and snowcover areal distribution provide useful information important in many snow-related management activities. This information finds use for:

(1) forecasting water supplies for: irrigation, livestock, power generation, fisheries, waterfowl, domestic uses, etc.;
(2) predicting flood-water peaks from snowmelt runoff;
(3) keeping snowdrifts from obstructing roads, feedlots, railways, and farm yards;
(4) accumulating clean snow in winter recreation areas;
(5) retaining snow over winter crops for low temperature protection; and
(6) estimating the soil water enrichments resulting from melting snow and used by dryland crops.

The ease with which snowpacks can be quantified varies widely from location to location. Unfortunately, accuracy in extrapolating snowpack measurements in expansive prairie environments is often difficult to obtain. For example, Canadian prairie snowcovers typically exhibit large areal variation. This variation results from the vast area within which the region's snowstorms may track and the dynamic, wind-swept nature of the region's prairie environments. The wind can blow an area bare of snow and redistribute the icy crystals into deep drifts located very close to bare areas. Extremes, ranging from shortfall to paralyzing blizzards, characterize the northern plains and prairies.

Presented at the 68th Annual Meeting of the Western Snow Conference, April 18-20, 2000, Port Angeles, Washington.

[1] Research Scientist, Semiarid Prairie Agricultural Research Centre, Agriculture and Argi-Food Canada, Swift Current, Saskatchewan

THE CANADIAN PRAIRIES

The Canadian Prairies include a loosely defined area approaching 700,000 km^2 (275,000 sq. miles). The region forms part of the North American mid-continental plains, sloping gradually east-northeastward from the Rocky Mountains to Hudson Bay and spanning parts of British Columbia, Alberta, Saskatchewan, and Manitoba (Figure 1). Landforms within the region are generally subdued, reflecting the effects of massive continental glaciation of the past. Topographic relief in excess of 80 m is rare and usually associated with water courses which currently carry mountain-borne, snowmelt-fed rivers across the Prairies or which once drained the receding continental glaciers.

Five major soil zones are recognized within the Prairie Provinces (Figure 1): Brown (Aridic Borolls), Dark Brown (Typic Borolls), Black (Udic Borolls), Dark Gray (Boralfic Cryoborolls), and Gray (Alfisols). Generally, reference to the Canadian Prairies includes all the areas included in the Brown and Dark Brown zones plus part or all of the Black. The Brown soil zone is the most arid, and the natural vegetation is typically a short grass mixed prairie. The Dark Brown zone is less arid, and the vegetation is of the mid-grass section of the mixed prairie. Black soils are typical of the fescue prairie-aspen grove (parkland) and true prairie grasslands.

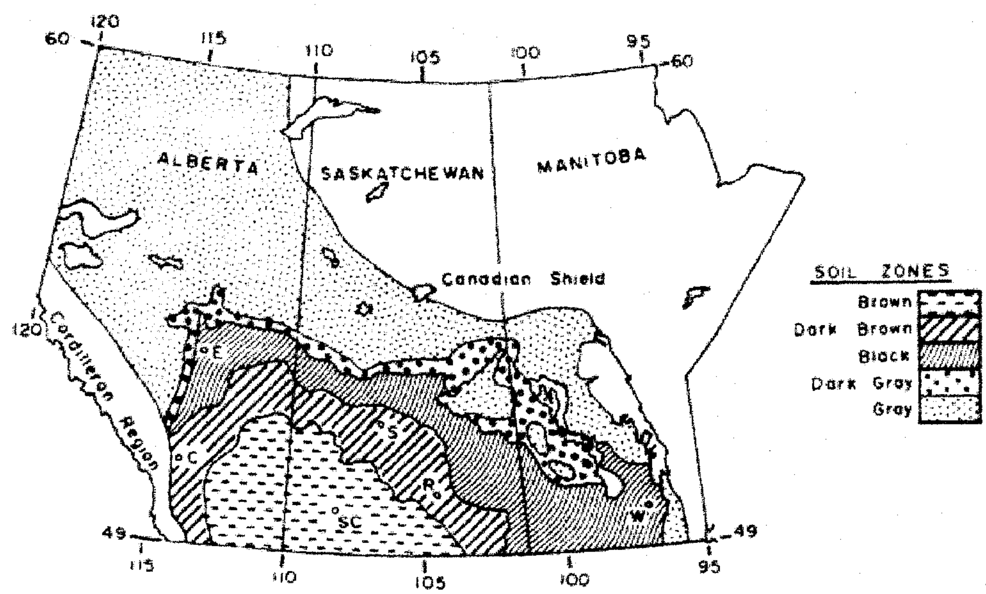


Figure 1. Generalized soils map of the Canadian Prairie Provinces (C=Calgary, E=Edmonton, S=Saskatoon, SC=Swift Current, R=Regina, W=Winnipeg)

Although the Canadian Prairie climate is classed as cool semiarid, it is well known for its extremes. The weather may range from hot to cold and from very dry to very wet. Winter air temperatures may drop well below -40°C and reach +35°C during the summer. Table 1 reveals the yearly fluctuations in mean precipitation and evaporation encountered at one location on the Prairies. Mean annual precipitation ranges from 300 mm in the Brown soil zone to 500 mm in the Black, with up to two-thirds falling as rain between April and November. During the summer, potential evaporation greatly exceeds precipitation, owing to ample solar radiation, warm temperatures, and the ever-characteristic prairie wind. Indeed, the wind contributes a major element to the dynamic nature of the climates on the Prairies, especially in winter.

Table 1. Accumulated daily precipitation and pan evaporation at the Agricultural Research Farm near Swift Current, Saskatchewan (Brown soil zone)

	Precipitation (mm) from standard gauges			Class A pan
		during previous	total during	Evaporation (mm)
Year	May-Aug.	winter and fall	water year	May-Aug.
1967	89	185	274	1087
1968	118	147	265	992
1969	135	200	335	1033
1970	245	170	415	958
1971	125	150	275	1104
1972	147	105	252	993
1973	81	180	261	1138
1974	257	202	459	911
1975	208	138	346	855
1976	229	177	406	1069
1977	224	66	290	878
1978	136	145	281	943
1979	166	164	330	948
1980	191	96	287	947
1981	212	143	355	952
1982	285	124	409	758
1983	205	143	348	1018
1984	119	83	202	1180
18-yr mean	176	145	321	987
+Long-term mean	209	151	360	988

+Period of 100 years for precipitation and 25 years for evaporation.

On average but not always, up to one-third of the annual precipitation falls as snow. A wide deviation in seasonal accumulation exists and stems from the region's size and dynamic climate. Storms which deposit 10 cm or more of snow typically occur only two to five times a season. Although usually wide-spread, these storms do not produce uniform snowfalls. They even miss some districts completely, because the fast-tracking, wind-driven storms rarely cover the entire region. High winds, blowing across the subdued, agricultural terrain, are responsible for considerable areal redistribution and sublimation of snowfall and snowcover. Snowpack accumulations also vary in response to an open-sky radiation and energy advection between shifting air masses. These may initiate thaw and melt at anytime, leading to snowpack losses through evaporation and meltwater releases.

The dynamic character of the prairie environment is reflected in the degree of snowcover permanency as described by McKay (1964) and depicted in Figure 2. Snowpacks over the Brown soil zone frequently disappear and reform in response to varying weather and chinook winds. East, north and northwestward of this zone snowcovers disappear less frequently until at the region's extremities they tend to persist throughout the winter.

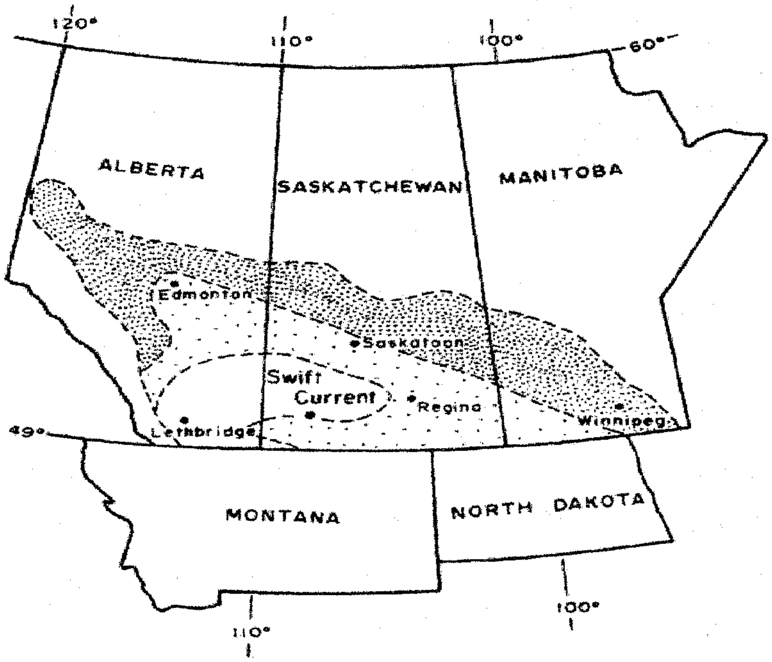


Figure 2. Generalized snowcover permanency during winter over the Canadian Prairies; degree of shading reflects permanency from a zone where snowcovers are frequently lost (white) to where they remain throughout the winter (dark).

ESTIMATING PRAIRIE SNOWPACKS

Snow Surveys

Scientific estimates of snowpack depth, liquid water equivalent and area coverage are needed to effectively manage the prairie snow resource. Accurate areal estimates are often difficult to obtain. The traditional snow survey using point measures obtained by weighing vertical snowpack cores encounters enormous sampling problems. If one assumes that each observation accurately describes the absolute water equivalent covering the immediate one square meter of land, a sample size of ten observations for every 1000 km^2 results in a 1/100,000,000 sample (10 m^2 per 1000 km^2). Areal measurements of prairie snowpacks by sensing the attenuation of terrestrial gamma radiation may ease the sampling task, but will not eliminate the low sampling ratio.

The snow water equivalent, WE, at a point, i, expressed per unit area is commonly computed as the product of the snow depth, d, and the specific gravity (density), f, of a vertically integrated snow column at the point:

$$WE_i = (f)_i\,(d)_i \quad [1].$$

The mean areal $\underline{WE}$ for any area of interest can be estimated by using an arithmetic average of n number of sample columns obtained throughout the area sampled:

$$\underline{WE} = \sum_{i=1}^{n} WE_i \quad [2].$$

The $\underline{WE}$ can also be estimated by using areal mean values for snowpack depth, $\underline{d}$, and specific gravity, $\underline{f}$, obtained from snow survey sampling of the pack to be estimated (Steppuhn, 1975):

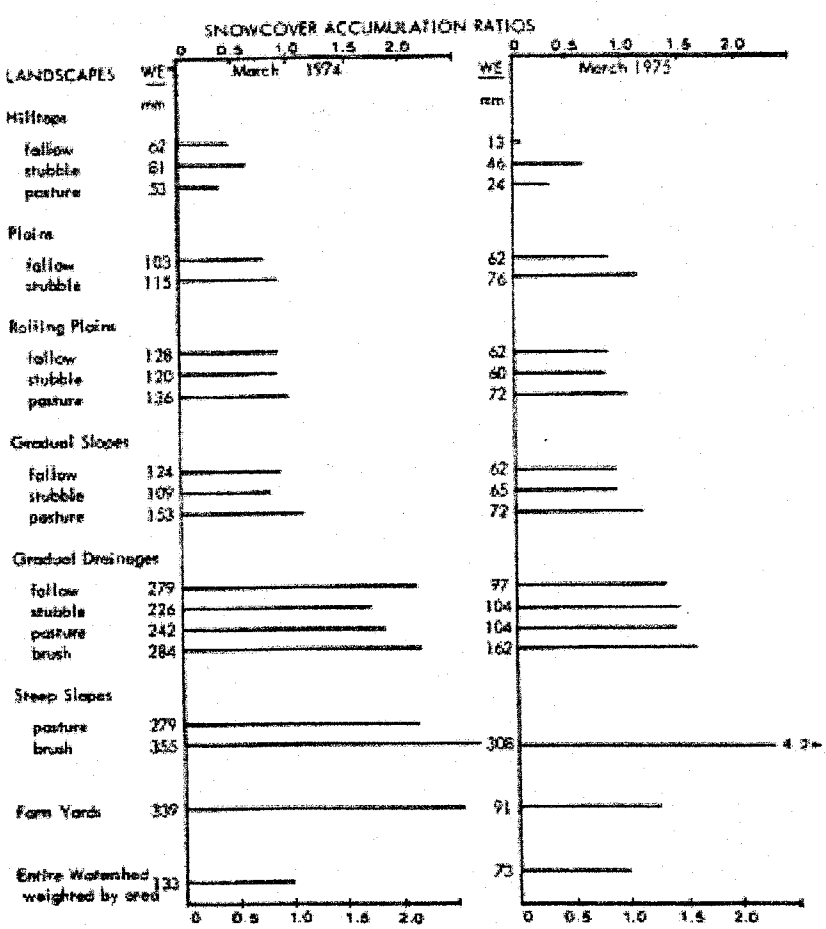
$$\underline{WE} = (\underline{f})\,(\underline{d}) + (r\,s_f\,s_d) \quad [3],$$

where the correlation coefficient, r, between f and d measurements forms a term together with the sample standard deviations, s, for f and d. Estimating WE by this component separation technique, Equation [3], reduces the total sampling effort. The areal variability associated with f is much less than it is for d, permitting a reduction in the number of labor-intensive snow cores required (Steppuhn, 1976). This leaves more time to increase the number of d-values obtained, providing better sampling precision for this more variable component of WE.

Sample stratification is another technique for increasing the accuracy in estimating WE. Kuz'min (1960) suggested stratifying snowpacks according to terrain, while Steppuhn and Dyck (1974) extended the concept to also include vegetative cover. Table 2 presents a stratification scheme used by Steppuhn (1976) to estimate areal WE in the Brown soil zone. The procedure allows snow surveyors to statistically sample snowpacks covering stratified land units classified by terrain and vegetation. Snowpack d and f are sampled separately within each landscape class. Sample means are calculated and used to compute WE estimates which include statistical measures of sampling precision.

The snowpack water equivalents reported in Table 2 show that prairie snowcovers preferentially accumulate in the lowlands, within farm yards, behind sharp changes in slope, and where shrub vegetation dominates. These data reflect the dynamic effect of wind and, to a lesser degree, melt on prairie snowpacks. They also demonstrate the value of stratification for quantifying wind-swept prairie snowcovers.

Table 2. Relative snowcover accumulation ratios according to landscape classes on March 1974 and March 1975 comprehensive snow surveys, Creighton Watershed, Saskatchewan



* WE = Actual Snow Water Equivalent

Precipitation Gauge Measurements

Most water stored in any seasonal snowpack on the Prairies originates as snowfall, accumulating over time under winter conditions. Consequently, measures of precipitation, especially when occurring as snowfall, have been used to estimate snowpack water equivalent. Equations based on the conservation of mass describe the relationship:

$$WE_t = \sum_{j=1}^{t} (W_j + C_j - E_j - I_j - R_j + B_j) \qquad [4],$$

where the snowpack water equivalent, WE_t, at time t equals the algebraic sum of snowfall water, W, condensation, C, evaporation, E, infiltration into the surface litter or soil, I, runoff, R, and the net mass of deposited or eroded wind-blown snow, B, accumulated by time increments j over the period t beginning on the day that the snowcover begins to accumulate. In dynamic, windy environments, Equation [4] could be evaluated for each snowcover landscape unit identified in the watershed.

The accuracy with which snowfall gauge measurements can estimate WE depends firstly on the precision associated with precipitation gauges used. Systematic under-catch is common. Goodison (1978) compared the snowfall quantities caught by various gauges to that accumulated on sheltered snow-boards. Figure 3 traces his comparisons for the Alter-shielded Universal gauge and the Nipher-shielded Canadian gauge over a range of windspeeds operating at orifice height. Although the Nipher gauge performs comparatively well, Goodison recommends corrections for wind when the data are used for absolute determinations.

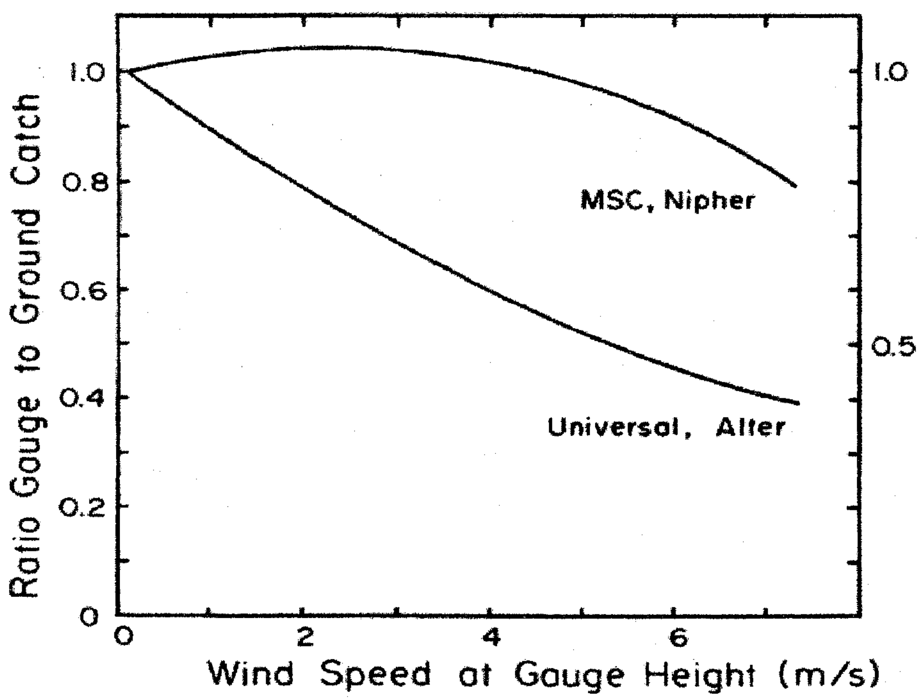


Figure 3. Relationship between snowfall catch and ground catch as a function of wind speed for Nipher MSC and Universal Alter-shielded snowfall gauges (Goodison, 1978).

The Agricultural Climatological Station on the Research Farm near Swift Current, Saskatchewan, forms part of the Federal Atmospheric Environment Service's (AES) Canadian network. The station is located 3 km southeast of the city on a wind-swept plain sloping gently (1% or less) to the north. The Nipher-shielded gauge complies with AES site requirements and is read manually at least once each day.

The agricultural fields and research plots surrounding the climate station accumulate snowpacks typical of open plains in a wind-blown prairie environment. The packs were measured for f and d 56 times during the 22 winters from 1965 through 1986. The snow surveys reflected the snowcovers existing on the fields and plots at the time of measurement and were located within 1 km of the Nipher gauge. Two types of vegetative covers were sampled: wheat stubble and cultivated summerfallow.

Liquid water volumes, W, from daily accumulations of snowfall caught in the Nipher gauge served as point measures of precipitation. These quantities, left uncorrected for wind, were summed by daily increments of j over the accumulation period, t, according to each snow survey date,

$\sum_{j=1}^{t} (W_j)$. A total of 56 summations were also available for comparison with the WE-values

calculated from the snow survey data applied to Equation [2].

Comparing Snowpack Estimates

The comparisons of WE from the snow surveys plotted as functions of accumulated W from the Nipher gauge are shown in Figure 4 for stubble, fallow, and stubble-fallow combined. Linear regressions based on the 56 data pairs resulted in r^2 values of 0.30, 0.42, and 0.41, respectively, for the three surface covers. In all comparisons, snowpack volumes estimated from the gauge exceed those actually measured on the ground. These results follow the basic premise of Equation [4], that snowpack volumes are subject to change, especially those associated with evaporation and meltwater release. Perhaps Nipher-based estimates would compare more closely with the ground snow surveys, if they were adjusted for these factors.

ADJUSTING PRECIPITATION GAUGE ESTIMATES

One of the daily observations required at Canadian AES Climatological Stations involves measuring the depth of the snowpack in level areas within and immediately surrounding the site containing the weather instruments. These depths are averaged and reported each day as "Depth-of-Snow-on-the-Ground." If these daily depths increase in tandem with increases in snowfall precipitation, new snow is being added to the pack. If depths decrease after having previously remained constant, the pack is increasing in density and/or losing snow-mass by meltwater runoff, subsurface infiltration, or evaporation as outlined in Equation [4]. Thus, these depth measurements offer an opportunity to adjust the accumulated Nipher gauge data for weather-induced snowpack changes.

The adjustment procedure requires daily-measured inputs of snowfall, W_j, and depth-of-snow-on-the-ground, h_j. These observations serve the calculation of a daily specific gravity (density) value, f'_j, and a daily-adjusted cumulative water equivalent, W'_j of a theoretical snowpack, whose dynamics are based on actual weather measurements. Depending on W_j, h_j, and f'_j, W'_j is used to adjust W_j for evaporation and meltwater releases. The procedure does not adjust the snowpack WE for condensation nor for net change due to erosion or deposition by wind-borne snow, nor does it separate melt losses into runoff or infiltration. Once the theoretical specific gravity for the day, f', is computed, the adjustment method requires three additional assumptions:

(1) that the maximum f-value for prairie packs equals 0.35 (or any other chosen value),
(2) that no significant change in snowpack mass occurs until the maximum density is reached (thus, evaporation without depth change is ignored), and
(3) that water in excess of maximum f' leaves the pack.

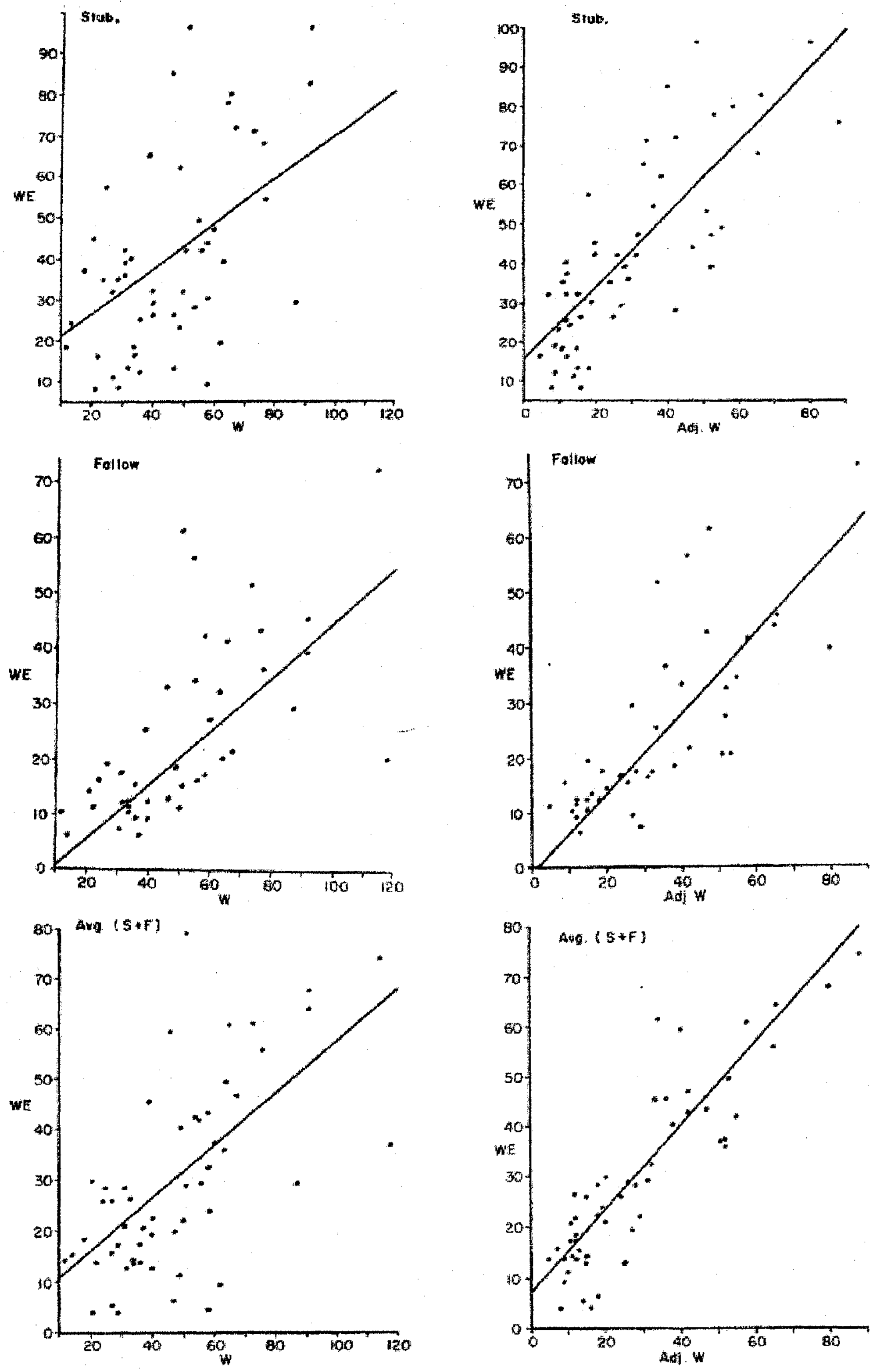


Figure 4. Plots of 56 mean ground-surveyed water equivalents, WE, for stubble (stub.), fallow, and combined (s & f) stubble and fallow land-covers, regressed with Nipher-shielded gauge-accumulated snowfall water, W, and W-adjusted for melt and evaporation, Research Farm, Swift Current, Saskatchewan

On any day, j, W_j can either equal or exceed zero, and the h_j measured that day will relate to the previous day's measurement, h_{j-1} in one of three ways: $h_{j-1} \leq h_j < h_{j-1}$. These equalities reflect sets of conditions under which the theoretical snowpack can respond. Each of these sets triggers specific equations for calculating W'_j and f'_j and reflects the snowpack physics indicated by the weather measurements. Depending on W_j and h_j, adjustments fall into one of three groups:

I. If $W_j > 0$ and $h_j > 0$ and $h_{j-1} \leq h_j < h_{j-1}$,

$$f'_j = W'_j / h_j \quad \text{and} \quad W'_j = W'_{j-1} + W_j \quad \text{and} \quad Adj.W_j = W_j \qquad [5]$$

Snowfall occurred, and the density remained below the threshold required for the pack to lose water.

II. If $W_j = 0$ and $h_j \geq h_{j-1}$,

$$f'_j = f'_{j-1} \quad \text{and} \quad W'_j = W'_{j-1} + W_j \quad \text{and} \quad Adj.W_j = 0 \qquad [6]$$

No new snow fell, and weather conditions allowed the snowpack density to remain below the threshold required for the pack to lose water.

III. If $W_j = 0$ and $h_j < h_{j-1}$,

$$f'_j = (h_{j-1} / h_j)\, f'_{j-1} \qquad [7]$$

No new snow fell, and weather conditions caused the snowpack to densify and perhaps lose water in proportion to decreases in measured h. Two sub-conditions further dictate additional applicable equations:

i) If $f'_j < 0.35$,

$$W'_j = W'_{j-1} + W_j \quad \text{and} \quad Adj.W_j = 0 \qquad [8]$$

ii) If $f'_j \geq 0.35$, set $f'_{j+1} = 0.35$

$$W'_j = W'_{j-1} - \{0.35\,(h_{j-1} - h_j)[1 - \frac{(f'_j - 0.35)}{(f'_j - f'_{j-1})}]\} \quad \text{and} \quad Adj.W_j = W'_j - W'_{j-1} \qquad [9],$$

where f'_j is not allowed to exceed a theoretical maximum, say 0.35, and from Equation [1], $0.35\,(h_{j-1} - h_j)$ equals the maximum potential volume of water which could have left the pack based on the measured change in snowpack depth. The term $1 - [(f'_{j-1} - 0.35)/(f'_j - f'_{j-1})]$ represents that fraction of the snowfall water associated with a change in depth in excess of densification to the maximum.

COMPARING ADJUSTED SNOWPACK ESTIMATES

The Research Farm snow survey data for 1965-86 were again plotted but this time as a function of the seasonally-accumulated Nipher gauge snowfall, W'_j, adjusted for evaporation and meltwater releases. The survey dates dictated the gauge summation periods. Graphs and linear regressions were again constructed for the three vegetation cover classes (Figure 4). Adjusting the accumulated snowfalls caught in the Nipher gauge for snowpack melt and evaporation using the site-depth procedure resulted in closer estimates of the measured snowpack WE values after this adjustment than before it (Table 3). Two-thirds or more of the variation in the snowcover WE could be explained by a linear regression with adjusted gauge-measured snowfall within a 5% error probability for a ±3.8 mm confidence about the mean. This correlation corroborates the use of adjusted snowfall precipitation measurements to estimate areal snowpack WE in a dynamic environment with reasonable accuracy.

Table 3. Linear regression statistics for predicting snowpack water equivalent as a function of accumulated Nipher gauge catch unadjusted, W, and adjusted, W', for melt and evaporation; Swift Current research fields in stubble and fallow, 56 points from the 1965-1986 snowcovers.

		Regression Statistic		
Vegetative Cover	Gauge Data	r^2	Std. Dev. (± mm)	Sample Size
Wheat Stubble	W	0.30	20	56
Wheat Stubble	W'	0.64	14	56
Fallow	W	0.42	14	56
Fallow	W'	0.66	10	56
Average of Stubble + Fallow	W	0.41	15	56
Average of Stubble + Fallow	W'	0.76	9	56

W = Accumulated snowfall water caught in a Nipher-shield precipitation gauge.
W'= W adjusted for snowpack evaporation and meltwater releases using climate station observations.
r^2= Coefficient of determination.
Std. Dev. = Standard deviation of the regression.

CONCLUSIONS

The quantification of the snowpack water equivalent covering landscapes on wind-swept prairies and plains can be eased by following three practices:
(1) Take all snowpack and climatological observations with the greatest of care and precision, because prairie snowpack accumulations are often small, causing errors to become large in percentage terms.
(2) To minimize the effects of wind and differential snowmelt, divide the watershed into snowcover landscape units and sample (or estimate) snowpack volumes according to the resulting stratification. This reduces the total variability and enhances accuracy in extrapolating estimates.
(3) Snowpack water equivalents may be estimated from Nipher-shielded precipitation gauge observations provided the accumulated value is adjusted for meltwater releases and evaporation. Relationships between gauge-measured snowfall and snowpack water equivalent should be established for each landscape type. This technique can be expected to give cumulative WE estimates within an error of ± 4 mm 95% of the time under typical conditions.

LITERATURE CITED

Goodison, B.E. 1978. Accuracy of Canadian Snowgauge measurements. J. Applied Meteorology 27:1542-1548.

Kuz'min, P.P. 1960. Snowcover and snow reserves. Gidrometeorologisheskoe Izdatel'skvo, Leningrad. [English Transl. by U.S. National Science Foundation, Washington, D.C., pp. 99-105.]

McKay, G.A. 1964. Relationships between snow survey and climatological measurements for the Canadian Great Plains. Proc. 32nd Annual Meeting, Western Snow Conf., pp. 9-18.

Steppuhn, H. 1975. Accuracy in estimating snowcover water equivalents. Proc. Canada Hydrology Symp.-75. Assoc. Comm. on Hydrology. National Research Council. pp. 36-41.

Steppuhn, H. 1976. Areal water equivalents for prairie snowcovers by centralized sampling. Proc. 44th Annual. Meeting, Western Snow Conf., pp. 63-68.

Steppuhn, H. and G.E. Dyck. 1974. Estimating true basin snowcover. Proc. Symp. on Advanced Concepts and Techniques in the Study of Snow and Ice Resources. U.S. National Academy of Sciences, pp. 314-328.